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A One-Dimensional Atmospheric Boundary Layer Model: Comparison with Observations

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Abstract

This report examines details of a one-dimensional (1D) atmospheric boundary layer model to establish the proper functioning of its soil, plant, and atmospheric physics. To achieve this goal, I inspect, repair, and modify a computer program that scientists at the Hebrew University, Department of Soil and Water Sciences, gave to me years ago. The computer program was exercised to determine if the model results are stable when initial conditions are changed and to determine whether the results are sensible and generally consistent with observed data. To show this, I present a time series of the modeled surface energy budget and modeled profiles of boundary layer wind speed, potential temperature, and specific humidity for daytime (atmospherically unstable conditions) and for nighttime (atmospherically stable conditions). I compare these results, wherever practical, with observed meteorological data. From these results, I infer how well the transfers of momentum, heat, and moisture from one model layer to the next are characterized. I also present root mean square error and d values, where d is an index of agreement, to summarize the model results and comparison with observed data. From the results, I find that the 1D model is functioning properly in solving for many parameter relationships and is as reliable as the earlier models of this type in predicting the general features of boundary layer development.

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1. Introduction

The atmospheric boundary layer is also called the friction layer. It extends from the earth's surface to the geostrophic wind (or gradient wind) level. Above this layer is the free atmosphere—where the frictional influence of the earth's surface is greatly diminished, allowing for an approximate balance between pressure gradient and Coriolis accelerations (Huschke, 1959). The daytime boundary layer is often observed to heights of 1 to 3 km above-ground level (agl), generating convective eddies (thermal updrafts and downdrafts) and relatively well-mixed profiles of wind speed, temperature, and moisture. In contrast, the nighttime boundary layer is characterized by a temperature inversion caused by a strong radiative cooling at the surface. The nighttime boundary layer can be at heights of 300 to 400 m agl. A low-level wind maximum or jet can sometimes develop in the nighttime boundary layer with faster, warmer air flow aloft, and slower, cooler air at the surface. This development of a low-level jet can promote even further cooling of the surface layer air, unless the inversion breaks down or overturns, which may occur because of increased wind shears or other larger-scale instabilities, perhaps also in combination with atmospheric waves (Businger, 1973).

In formulating an atmospheric boundary layer study, one needs a useful model calculation. That is, are the soil, plant, and atmospheric physics in the model functioning correctly and reliably? To achieve this goal, one needs to study the details of a model program to determine if the model results are stable and whether the results are sensible, i.e., generally consistent in comparison with observed data.

In this report, I use a computer model (see sect. 2) that scientists at the Hebrew University, Department of Soil and Water Sciences, gave to me several years ago. To test the model, I generate a time series of the surface energy budget and profiles of boundary layer wind speed, potential temperature, and specific humidity for daytime (atmospherically unstable conditions) and for nighttime (atmospherically stable conditions). I compare these model results, wherever practical, with observations from days 33 and 34 of the Wangara experiment (Clarke et al, 1971). From these results, I examine how the model characterizes transfers of momentum, heat, and moisture. I attempt to quantify these results by preparing root mean square error (rmse) and d values, where d is an index of agreement (described in sect. 5), to summarize the model comparison with observed data.

2. Model Description

The one-dimensional soil, plant, and atmospheric model used in this report has been documented previously in Pielke and Mahrer (1975), McNider and Pielke (1981), and Avissar and Mahrer (1988). It is a first-order closure model to calculate the transfer of momentum, heat, and moisture at the surface and aloft. It uses an implicit finite difference scheme to integrate the boundary layer and soil diffusion equations. It contains a complete model of the surface energy budget and a time-dependent calculation of the height of the daytime planetary boundary layer. I added the formulation suggested by Smeda (1979) for the time-dependent calculation of the height of the nighttime stable layer. The model surface layer turbulence scaling is as described by Zilitinkevich (1970) and Businger et al (1971).

As initial input, the model requires day of the year and time of day data, latitude and longitude, fraction of sky cloudiness, and ground-cover data (i.e., canopy height, leaf area index, surface reflectivity (albedo), thermal emissivity, and surface roughness). Typical values for albedo and surface roughness have been documented by Hansen (1993a, 1993b). The model computes soil properties (e.g., hydraulic and thermal conductivity and soil-specific heat capacity) from inputs of soil water content, porosity, texture (i.e., proportion of sand, clay, and organic matter), and bulk density. Other model constants include subsoil properties, such as plant root density and distribution. Also, the model requires an initial profile of wind speed, air temperature, and specific humidity (atmospheric water vapor content) from the ground level to the top of the model. Model parameters for this report are summarized in table 1.

The model equations for the time-dependent calculation of the winds (u and v components), potential temperature (θ), and specific humidity (q) over flat earth can be expressed as

$$\frac{\partial \overline{u}}{\partial t} = f\left(\overline{v} - v_g\right) + \frac{\partial}{\partial z}\left(K_m \frac{\partial \overline{u}}{\partial z}\right), \tag{1}$$

$$\frac{\partial \overline{v}}{\partial t} = f\left(u_g - \overline{u}\right) + \frac{\partial}{\partial z}\left(K_m \frac{\partial \overline{v}}{\partial z}\right), \tag{2}$$

$$\frac{\partial \overline{\theta}}{\partial t} = \frac{\partial}{\partial z}\left(K_h \frac{\partial \overline{\theta}}{\partial z}\right), \text{ and} \tag{3}$$

$$\frac{\partial \overline{q}}{\partial t} = \frac{\partial}{\partial z}\left(K_q \frac{\partial \overline{q}}{\partial z}\right), \tag{4}$$

where f denotes the Coriolis parameter (the deflecting force caused by the earth's rotation acting upon moving air), the subscript g refers to the geostrophic wind (such that the first term on the right contains the pressure gradient acceleration), K_m denotes the eddy transfer coefficient for

Table 1. Boundary layer model parameters.

Parameter	Value
Number of vertical levels	32 (2, 10, 50, 100, 150, 200, 250, 300, 350, 400, 450, 500, 550, 600, 650, 700, 750, 800, 850, 900, 950, 1000, 1100, 1200, 1300, 1400, 1500, 1600, 1700, 1800, 1900, 2000 m)
Latitude, longitude	34.50 S, 144.93 E
Surface roughness	0.0045 m
Vegetation (sparse)	0.01 m
Surface albedo	0.20
Surface emissivity	0.98
Soil water content	0.08 m^3/m^3
Average soil density	1600 kg/m^3
Soil texture (% sand, % clay, % organic)	28.0, 70.0, 2.0
Day, month, year	16–17 August 1967
Geostrophic wind (u_g, v_g)	−5.34 m/s, −0.43 m/s
Initial time	09 lt
Time step	10 s

momentum, and $K_h = K_q$ denotes the transfer coefficients for heat and moisture. In the surface layer, K_m and K_h are calculated as $K_m = \mathrm{k}u_* z/\phi_m$ and $K_h = \mathrm{k}u_* z/\phi_h$, respectively, where k is Karman's constant (= 0.4); z is height above ground level (in meters); u_* is the friction velocity (in units of $\mathrm{m/s}^{-1}$), which relates to surface stress; and the ϕ_m and ϕ_h are nondimensional lapse-rate functions, which account for surface-layer stabilities other than neutral. A list of symbol definitions is provided in the appendix.

In the unstable daytime boundary layer, the model derives the transfer coefficients as a function of height, as suggested by O'Brien (1970). This function can be expressed as

$$K_z = K_{z_i} + \left(\frac{(z_i - z)^2}{(z_i - z_s)^2}\right)\left\{K_{z_s} - K_{z_i} + (z - z_s) \times \left[\frac{\partial K_{z_s}}{\partial z} + \frac{2\left(K_{z_s} - K_{z_i}\right)}{(z_i - z_s)}\right]\right\}, \tag{5}$$

where z is height above ground level and z_i and z_s refer to the heights of the top of the surface layer and top of the boundary layer, respectively. The profile function is applied similarly for the momentum, heat, and moisture coefficients. The model sets $z_s = 0.04z_i$, $K_{z_i} = 1.0$, and $K_z = 1.0$ for $z \geq z_i$. Figure 1 illustrates the geometry (i.e., vertical levels) of the model.

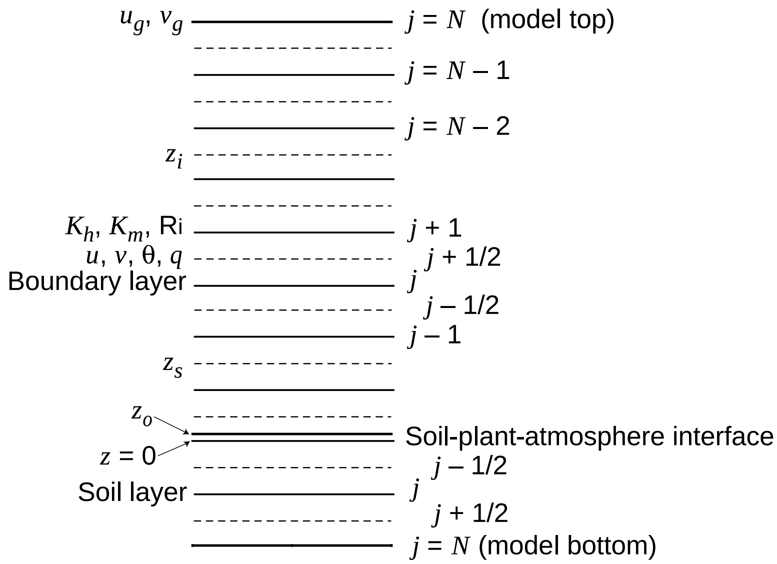


Figure 1. Sketch of geometry (i.e., vertical levels) of model.

The growth of the daytime planetary boundary layer is calculated with an expression derived by Deardorff (1974), which can be written as

$$\frac{dz_i}{dt} - w_{zi} = \frac{1.8\left(w_*^3 + 1.1u_*^3 - 3.3u_*^2 f z_i\right)}{g\left(\frac{z_i^2}{\theta_s}\right)\left(\frac{\partial\theta^+}{\partial z}\right) + 9w_*^2 + 7.2u_*^2}, \tag{6}$$

where w_{zi} is the vertical velocity at z_i (w_{zi} is assumed to be negligible or equal to zero, since time-dependent calculations of vertical velocities are not explicitly derived), $\partial\theta^+/\partial z$ is the vertical gradient of potential temperature in the stable air immediately above the boundary-layer top, θ_s is the potential temperature at the top of the surface layer z_s, and $w_* = \left[(-g/\theta)\, u_*\theta_* z_i\right]^{1/3}$ is the vertical velocity scaling variable, an implicit calculation of buoyancy, where g is acceleration caused by gravity, and $\theta_* = \frac{kz}{\phi_h}\frac{\partial\theta}{\partial z}$, the potential temperature scaling constant.

During the nighttime, when the surface layer is stable (i.e., $\theta_* > 0$), the model derives eddy transfer coefficients above the surface layer,* as suggested by Blackadar (1979), i.e.,

$$K_m(z) = K_h(z) = \begin{cases} sl^2\left(1 - 18\mathrm{Ri}\right)^{0.5}, \mathrm{Ri} < 0 \\ sl^2\left(\mathrm{Ri}_{\mathrm{crit}} - \mathrm{Ri}\right)/\mathrm{Ri}_{\mathrm{crit}}, 0 < \mathrm{Ri} \le \mathrm{Ri}_{\mathrm{crit}} \\ 0, \mathrm{Ri} > \mathrm{Ri}_{\mathrm{crit}} \end{cases}, \tag{7}$$

where s is the local wind shear, $s = \sqrt{(\partial u/\partial z)^2 + (\partial v/\partial z)^2}$, and Ri is the ratio of thermal to mechanical (wind shear) production turbulent energy called the Richardson number, so that $\mathrm{Ri} = \frac{g}{\theta}\frac{\partial\theta}{\partial z}/\left((\partial u/\partial z)^2 + (\partial v/\partial z)^2\right)$.

*Note: $z_s = 0.15\ z_i$ while $\theta_* > 0$.

$\mathrm{Ri}_{\mathrm{crit}}$ is the limiting value of the Richardson number, and it is often assumed that $\mathrm{Ri}_{\mathrm{crit}} = 0.25$, even though $\mathrm{Ri}_{\mathrm{crit}} = 1.0$ can sometimes be a useful approximation (Avissar and Mahrer, 1988). The length l (in meters) is generally thought of as the width of turbulence and can be characterized by the formulation reported in Blackadar (1979) for $z > z_s$ as

$$l = kz\left(1 + \frac{kz}{0.0063u_*/f}\right)^{-1}. \tag{8}$$

The height of the nighttime planetary boundary layer is calculated with the formulation suggested by Smeda (1979), which can be expressed as

$$\frac{dz_i}{dt} = 0.06\frac{u_*^2}{z_i f}\left[1 - \left(\frac{3.3z_i f}{u_*}\right)^3\right]. \tag{9}$$

The surface energy budget is applied to the soil, the plant canopy and the air throughout the canopy, and the thin layer of air that extends above the canopy top (Avissar et al, 1986). Air temperature and water vapor content for each layer are derived because of energy and water vapor transfers within the system. The net radiative flux is approximated as functions of transmission, albedo, leaf area, and soil wetness. The model radiation and energy budget as described in Pielke (1984) are

$$(1 - A)R_{s\downarrow} + R_{L\downarrow} - R_{L\uparrow} - \rho c_p u_* \theta_* - \rho L_v u_* q_* + Q_s = F, \tag{10}$$

where A is surface reflectivity (albedo); $R_{s\downarrow}$, $R_{L\downarrow}$, and $R_{L\uparrow}$ are the incoming solar, incoming long-wave, and outgoing long-wave radiative fluxes, respectively; ρ is air density; c_p is the specific heat of air at constant pressure; u_*, θ_*, and q_* are the surface-layer turbulence scaling parameters for wind speed, temperature, and moisture, in that order; L_v is the heat of transformation for water vapor; Q_s is the soil heat flux; and F is the function applied in solving for the surface temperature θ_{sfc}. Using the Newton-Raphson iterative algorithm, I can approximate each successive estimate of the surface temperature as

$$\theta_{\mathrm{sfc}}{}^{(m+1)} = \theta_{\mathrm{sfc}}{}^{(m)} - \frac{F}{F'}, \tag{11}$$

where $F' = \frac{\partial F}{\partial \theta_{\mathrm{sfc}}}$.

The model equations for the soil layer and plant canopy are described in much greater detail by Avissar and Mahrer (1988), with one exception. The model formulation for downward long-wave radiation is instead based on the empirical relationship reported by Paltridge and Platt (1976), which can be expressed as

$$R_{L\downarrow} = -170.9 + 1.195\,\sigma T_r^4 + 0.3\,\mathrm{cld}\varepsilon_c\,\sigma T_c^4, \tag{12}$$

where $\sigma = 5.6697 \times 10^{-8}$ W/m^2 deg^{-4}, T_r is the reference level (~2 m) temperature in Kelvin, cld is the cloud amount, ε_c is the emissivity of the

cloud base, and T_c is the temperature of the cloud base in Kelvin. This formulation is much simpler to apply than computing the upward and downward long-wave radiation according to the concentrations, path lengths, and emissivities for water vapor and carbon dioxide. Also, this expression was used successfully in two previous studies (Rachele and Tunick, 1994; Tunick et al, 1994).

3. Numerical Methods and Boundary Conditions

The numerical integration of equations (1) to (4) is achieved with the use of a generalized form of the Crank-Nicholson implicit finite difference scheme (Paegle et al, 1976) to solve for the vertical transfer of momentum, heat, and moisture in the atmosphere. The diffusion terms in equations (1) to (4) can be expressed in terms of both a current τ and future $\tau + 1$ timestep:

$$\frac{\phi^{\tau+1} - \phi^{\tau}}{\Delta t} = \left[\begin{array}{l} K_{j+1/2} \frac{\beta_\tau \left(\phi_{j+1}^{\tau} - \phi_j^{\tau}\right) + \beta_{\tau+1}\left(\phi_{j+1}^{\tau+1} - \phi_j^{\tau+1}\right)}{\Delta z_{j+1/2}} \\ -K_{j-1/2} \frac{\beta_\tau \left(\phi_j^{\tau} - \phi_{j-1}^{\tau}\right) + \beta_{\tau+1}\left(\phi_j^{\tau+1} - \phi_{j-1}^{\tau+1}\right)}{\Delta z_{j-1/2}} \end{array} \right], \tag{13}$$

where $\beta_\tau + \beta_{\tau+1} = 1$, $\Delta z_j = z_{j+1/2} - z_{j-1/2}$, $\Delta z_{j+1} = z_{j+1} - z_j$, and $\Delta z_{j-1} = z_j - z_{j-1}$. In this study, $\beta_\tau = 0.75$ and $\beta_{\tau+1} = 0.25$ are the weights to the current and future contributions to the numerical approximation (Pielke, 1984). Equation (13) can be rewritten as $a_j \phi_{j-1} + b_j \phi_j + c_j \phi_{j+1} = d_j$ such that

$$\begin{aligned} &- \frac{\beta_{\tau+1} K_{j-1/2} \Delta t}{\Delta z_j \Delta z_{j-1/2}} \phi_{j-1}^{\tau+1} + \left[1 + \frac{\beta_{\tau+1} K_{j+1/2} \Delta t}{\Delta z_j \Delta z_{j+1/2}} + \frac{\beta_{\tau+1} K_{j-1/2} \Delta t}{\Delta z_j \Delta z_{j-1/2}} \right] \phi_j^{\tau+1} - \frac{\beta_{\tau+1} K_{j+1/2} \Delta t}{\Delta z_j \Delta z_{j+1/2}} \phi_{j+1}^{\tau+1} \\ &= \phi_j^{\tau} + \left[\frac{\beta_\tau K_{j+1/2} \Delta t \left(\phi_{j+1}^{\tau} - \phi_j^{\tau} \right)}{\Delta z_j \Delta z_{j+1/2}} + \frac{\beta_\tau K_{j-1/2} \Delta t \left(\phi_j^{\tau} - \phi_{j-1}^{\tau} \right)}{\Delta z_j \Delta z_{j-1/2}} \right]. \end{aligned} \tag{14}$$

The model solves equation (14) using the coefficients a_j, b_j, c_j, and d_j and the algorithm described by Ahlberg et al (1967), which can be expressed as

$$\phi_j = x_j + y_j \phi_{j+1}, \tag{15}$$

where $x_j = (d_j - a_j x_{j-1})/p_j$, $p_j = b_j + a_j y_{j-1}$, and $y_j = -c_j/p_j$, and where $x_0 = y_0 = 0$. When the pressure gradient and Coriolis accelerations in equations (1) and (2) are included, the coefficient d_j (i.e., the right-hand side of eq. (14)) becomes $d_j = d_j + f(\overline{v} - v_g)\Delta t$ and $d_j = d_j + f(u_g - \bar{u})\Delta t$, respectively.

The boundary condition applied at the top of the model, i.e., $j = N$, for this solution is $\phi_N = 0.75\phi_{N-1} + 0.25\phi_{N-2}$, so that the variables near the top of the model are not fixed (i.e., not held to their initial values). This condition appears to behave well.

4. Days 33 and 34 Observations

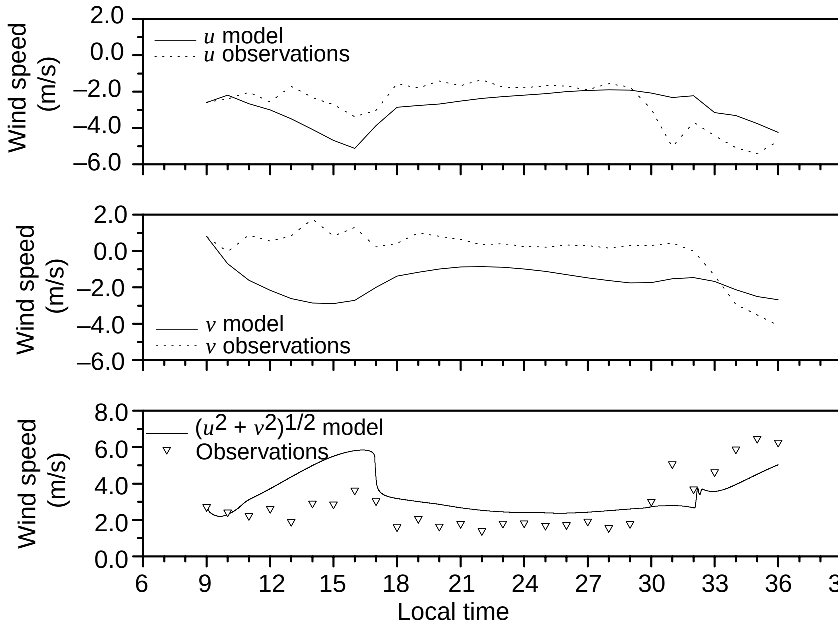
The Wangara experiment data were collected over a large flat area ($\sim$60 $\times$ 60 km), consisting mainly of thin low bushes and desert grasses. The field study was conducted near the town of Hay, New South Wales, Australia (34.50 S, 144.93 E), during the 1967 southern hemisphere winter (Clarke et al, 1971). Five types of data were collected: (1) hourly meteorological tower measurements of wind speed, specific humidity, and temperature; (2) hourly surface measurements of the net radiative and soil heat flux; (3) measurements of wind speed and wind direction from 0 to 2 km agl obtained by hourly pibal (pilot balloon) flights; (4) measurements of pressure, temperature, and mixing ratio (weight of water vapor/weight of dry air) to a height of 2 km, taken by radiosonde (instrumented balloon sonde) flights at 3-hour intervals; and (5) reports of fractional, low, high, and total cloud cover (including cloud-type descriptors).

These data are sufficent to test the functioning of boundary layer models (e.g., Pielke and Mahrer, 1975; Yamada and Mellor, 1975; McNider and Pielke, 1981). Of the 1050 published reports (profiles), I used those data for the 27-hour period from 09 lt (local time) on 16 August 1967 to 12 lt on 17 August 1967. This time period was characterized mainly by clear skies, relatively light daytime wind speeds, dry soil, and strong surface-based convective heating. In contrast, the nighttime boundary layer winds over this time period were quite strong, and the surface layer cooled significantly.

5. Model Results

Figure 2 gives the time series of the modeled u and v components of wind speed and the time series of the total wind, $\sqrt{u^2+v^2}$, at 2 m agl, in comparison with days 33 and 34 observations. The observed wind speeds are shown as mostly overpredicted, especially through the period 12 to 18 lt. As a result, the values of the index of agreement* calculated for these data are about 50 to 60 percent (see table 2). These comparisons might have been improved if adjustments to the geostrophic wind and thermal wind (i.e., u_g, v_g, $\partial u_g/\partial z$, and $\partial v_g/\partial z$) were included in the calculation. I use $z_0 = 0.45$ cm based on modeling the surface energy budget over a barren field (Tunick et al, 1994). Several others, however, have used values of z_0 other than 0.45 cm; for example, Pielke and Mahrer (1975) used $z_0 = 0.12$ cm, as reported on p. 21 of Clarke et al (1971), and McNider and Pielke (1981) used $z_0 = 1.0$ cm, based on the recommendations of Yamada and Mellor (1975) and others that 0.12 cm was too small for use with the surface-layer flux formulations in their models.

Figure 2. Modeled u and v components of wind speed and total wind, $\sqrt{u^2+v^2}$, at 2 m agl, compared with days 33 and 34 observations.



*The index of agreement d, as suggested by Willmott (1981), is calculated as

$$d = 1 - \sum_{i=1}^{n}(M_i - O_i)^2 \Big/ \sum_{i=1}^{n}\left[\left|(M_i-\bar{O}_i)\right| + \left|(O_i - \bar{O}_i)\right|\right]^2,$$

where M_i are the modeled data, O_i are the observed data, and the overbar corresponds to the mean. The nature of the index of agreement is a simple scale 0 to 100 percent, such that $d \to 1$ as the model predictions improve. In contrast, the rmse is calculated in the same units as the variable, i.e., rmse $= \left[N^{-1}\sum_{i=1}^{N}(M_i - O_i)^2\right]^{1/2}$.

Table 2. Summary of model results in comparison with observed data.

Model parameter	rmse	d	n
Total wind speed @ 2 m agl, m/s	1.749	0.592	28
u = wind speed @ 2 m agl, m/s	1.187	0.616	28
v = wind speed @ 2 m agl, m/s	2.148	0.480	28
Temperature @ 2 m agl, K	0.627	0.966	28
Net radiative flux, W/m^2	25.098	0.995	27
Soil heat flux, W/m^2	30.273	0.955	27
Boundary layer height, m	218.495	0.888	7
Wind speed profiles, m/s (12, 15, 18, 21 lt)	2.838	0.601	120
Wind speed profiles, m/s (24, 03, 06, 09 lt)	5.010	0.372	120
Potential temperature profiles, K (12, 15, 18, 21 lt)	0.435	0.992	120
Specific humidity profiles, g/kg (12, 15, 18, 21 lt)	0.209	0.988	120

Figure 3 gives the time series of the surface and 2-m air temperatures. The modeled surface temperatures θ_{sfc} are shown in comparison with results from an earlier modeling study reported by McNider and Pielke (1981). The modeled air temperatures at 2 m, $\theta_{2\mathrm{m}}$, are shown in comparison with the days 33 and 34 observations. The results are in very good agreement with these data, especially for the daytime hours. They imply that the model's surface energy budget calculations (shown in fig. 4) are functioning properly. Index of agreement d and rmse values for these data as well as the surface energy budget are given in table 2. The discrepancies between modeled and observed data at nighttime, however, may be related to initial values of surface roughness and to certain initial soil properties, such as soil water content.

Figure 5 gives the modeled height of the planetary boundary layer. The model formulations for z_i, which are based on calculations of the surface-layer turbulence scaling parameters, appear to be in fairly good agreement, i.e., $d \approx 90$ percent, with the observations, except perhaps during the hours immediately following sunset. The observations are few because they were estimated from days 33 and 34 profiles of θ and q, taken at 3-hour intervals (see Melgarejo and Deardorff, 1974, and Pielke and Mahrer, 1975).

Figure 6 gives the modeled boundary layer wind speed profiles at 3-hour intervals in comparison with days 33 and 34 observations. During the daytime, i.e., 12 to 18 lt, the profile average wind speeds (both modeled and observed) are shown to increase through the layer $z \leq 1300$ m. Similarly, the modeled and observed profiles of wind speed at night, i.e., 24 to 06 lt, show the development of a low-level jet in the layer $100 \leq z \leq 400$ m. However, agreement between the modeled and observed profiles for these data is generally poor (i.e., $d \approx 60\%$, daytime, and $d \approx 37\%$, nighttime), particularly through the upper layers and near the model's top. From this result, I suspect that either the upper boundary condition for equations (1) and (2)

Figure 3. Modeled surface ($\theta_{\rm sfc}$) temperatures compared with results from an earlier modeling study reported by McNider and Pielke (1981) and modeled 2-m ($\theta_{\rm 2m}$) air temperatures compared with days 33 and 34 observations.

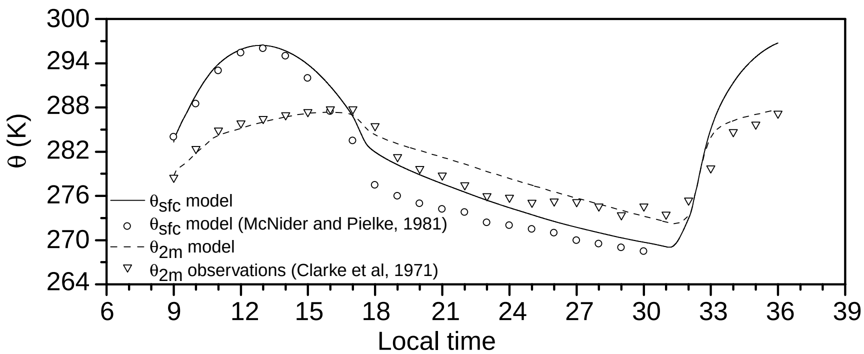


Figure 4. Modeled surface energy budget compared with days 33 and 34 observations of net radiative and soil heat flux.

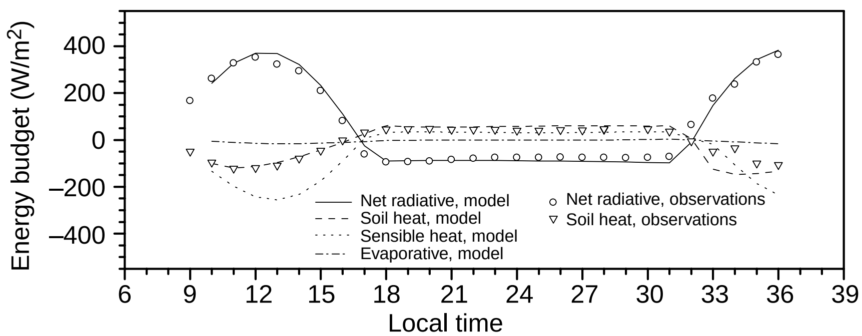
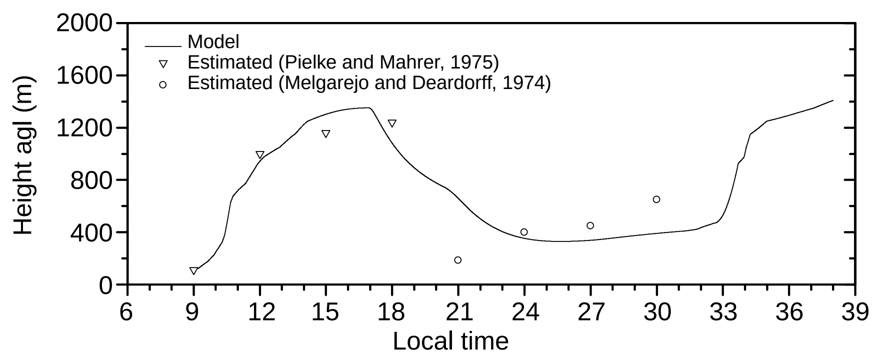


Figure 5. Modeled height of planetary boundary layer compared with days 33 and 34 observations.



does not apply and/or the observed profile data reflect changes (over time) in the geostrophic wind (see table 3), an effect that was not included in the calculation. This can be addressed in a separate study.

Figure 7 gives the modeled potential temperature and specific humidity profiles at 3-hour intervals in comparison with days 33 and 34 observations. The potential temperature profiles are well mixed vertically through the convective layer, $z \leq 1300$ m, increasing (on the average) over the daytime hours. The specific humidity profiles are also well mixed through the convective layer and are decreasing (on the average) over the daytime hours. The index of agreement between these modeled and observed data is $d \approx 0.99$. In part, this high value of agreement is because there is little or no advection of heat or moisture over time in the upper model layers. For layer $z \leq z_i$, I can infer that the eddy transfer (eddy diffusion) coefficients used in the model are formulated reasonably well. Figure 8 summarizes these comparisons of the modeled results to observed data.

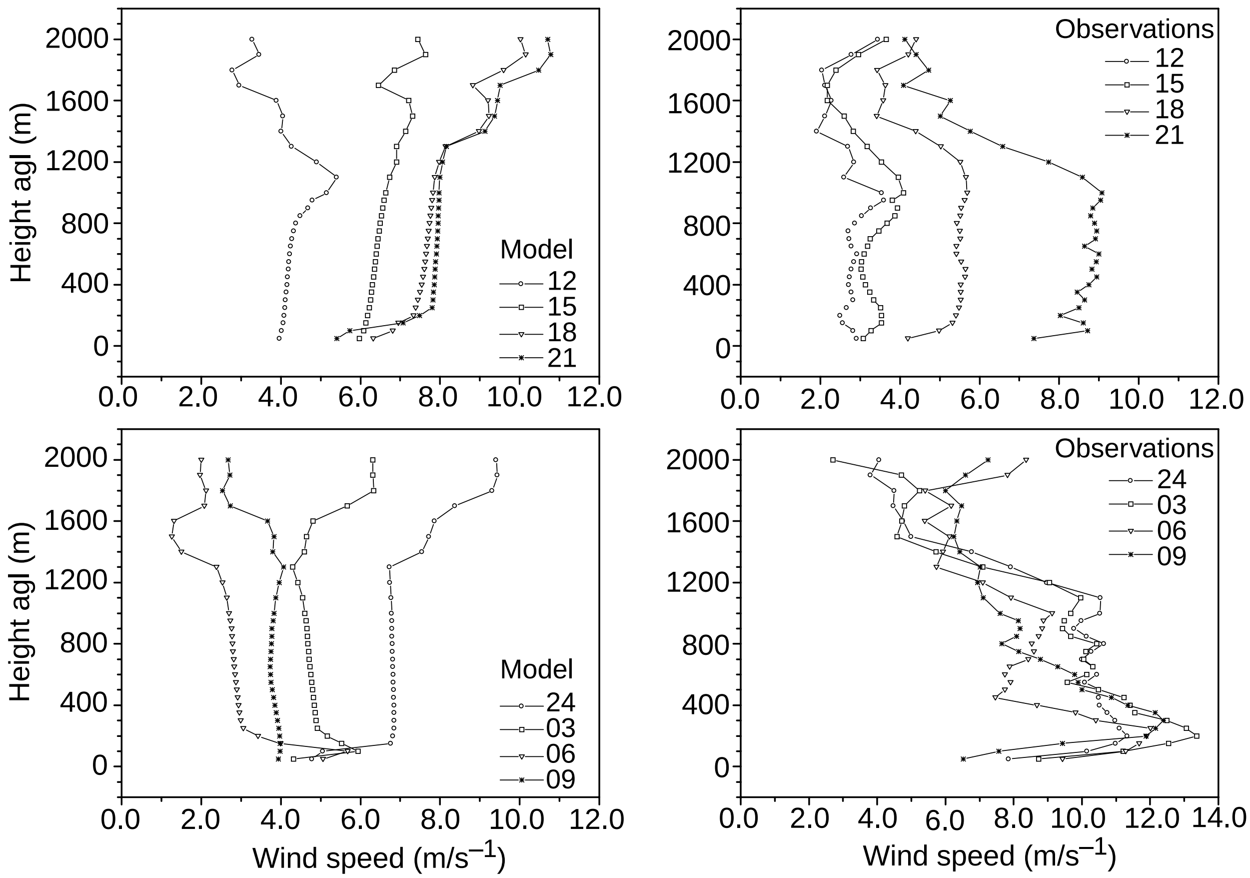


Figure 6. Model boundary layer wind speed profiles at 3-hour intervals compared with days 33 and 34 observations.

Table 3. Days 33 and 34 geostrophic wind data.

Local time	u_g (m/s)	v_g (m/s)	Total wind (m/s)	Direction (°)
09	−5.34	−0.43	5.36	85.4
12	−5.56	1.27	5.70	102.9
15	−6.20	0.98	6.28	99.0
18	−6.42	−0.72	6.46	83.6
21	−5.99	−1.93	6.29	72.1
24	−6.93	−2.86	7.50	67.6
03 (27)	−8.02	−3.13	8.61	68.9
06 (30)	−7.32	−4.97	8.85	55.8
09 (33)	−7.60	−4.72	8.95	58.2
12 (36)	−7.95	−3.43	8.66	66.7

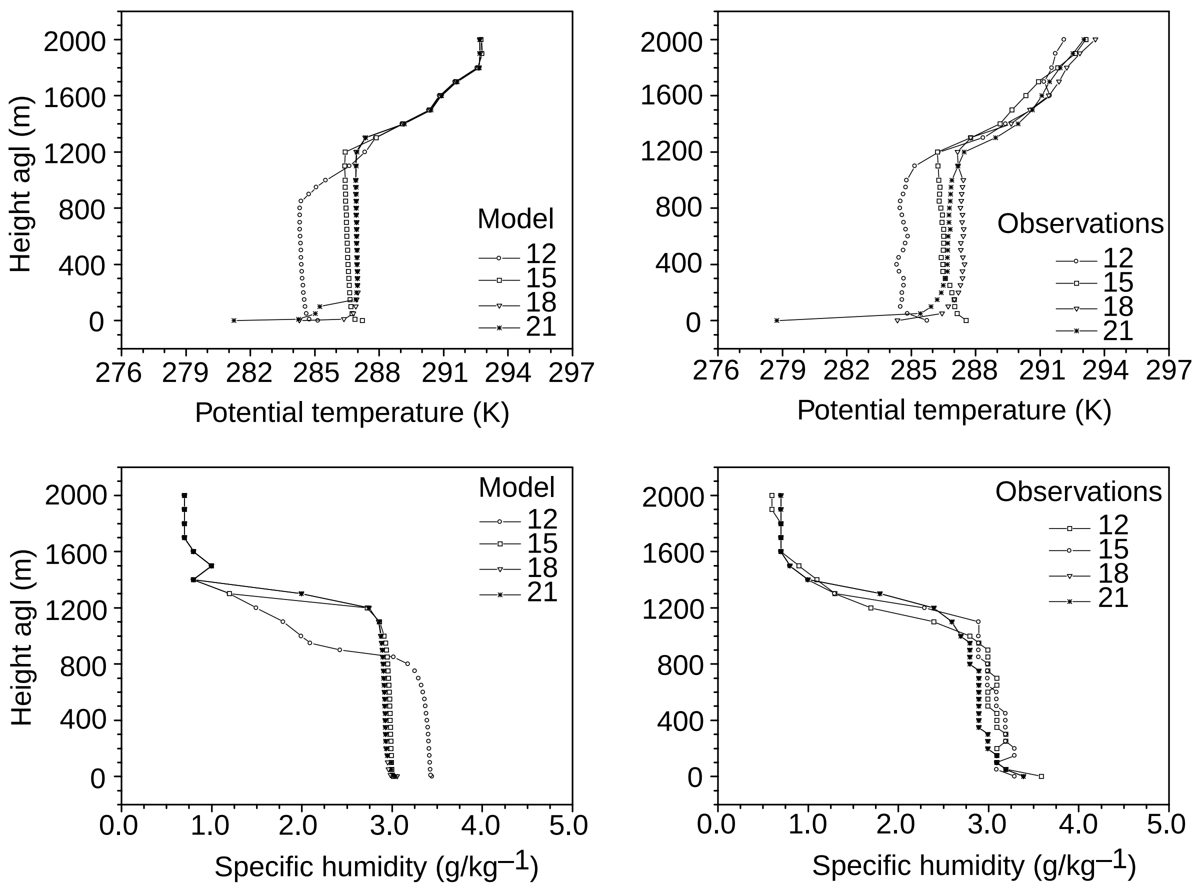


Figure 7. Modeled potential temperature and specific humidity profiles at 3-hour intervals compared with days 33 and 34 observations.

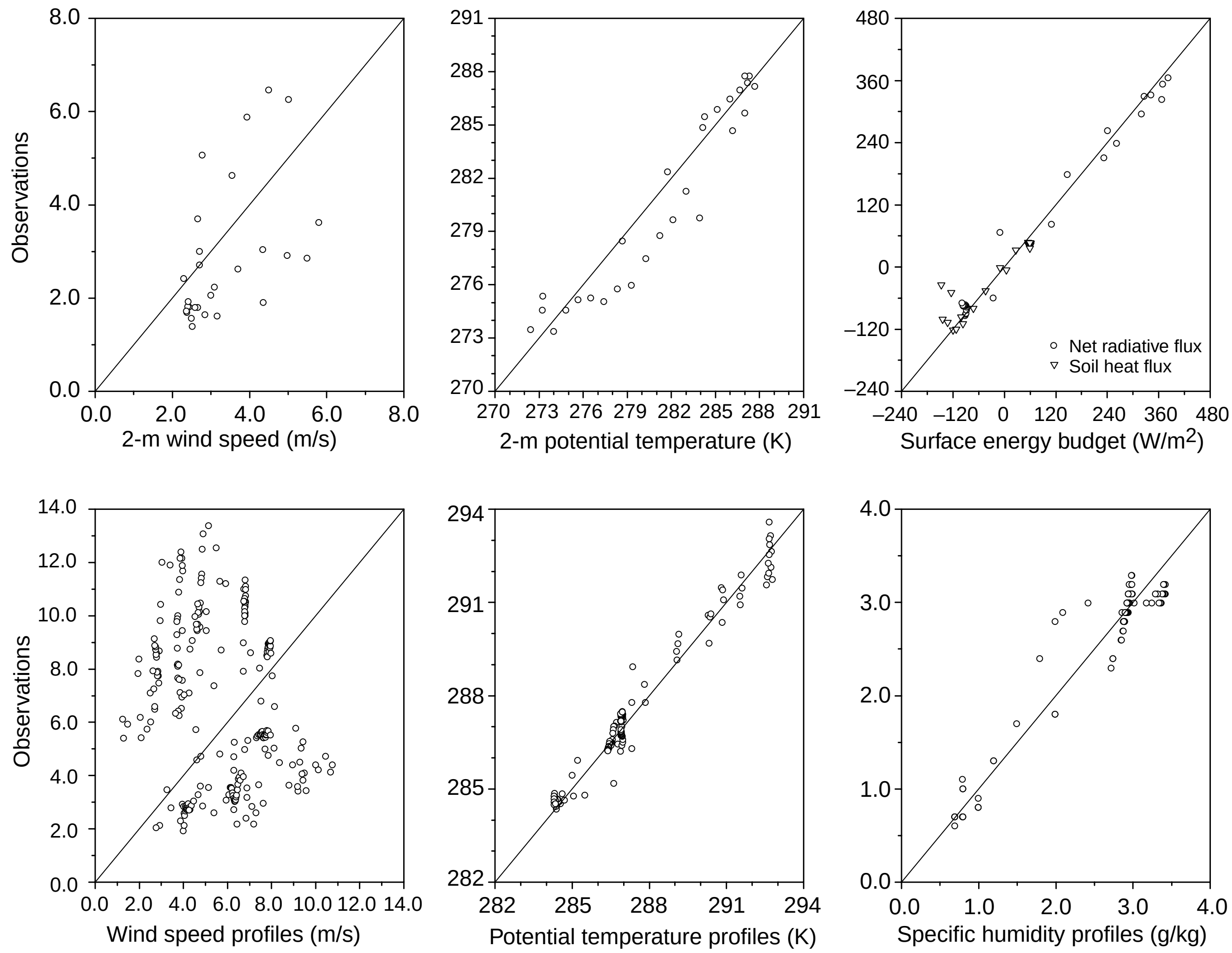


Figure 8. Scatter plot comparisons of modeled results to observed data.

6. Conclusion

For this report, I studied and modified an atmospheric boundary layer computer program. I ran the computer program, making changes to various sets of initial conditions. From this effort, I found that the one-dimensional model functions correctly and appears to be as reliable as any of the earlier models of this type (e.g., Pielke and Mahrer (1975) and McNider and Pielke (1981)) in predicting the general features of boundary layer development. I hope to use this work toward conducting an additional study on boundary layer wind shears under nighttime stable conditions.

I found this work to be very challenging. In particular, rendering the Ahlberg et al (1967) algorithm from the computer code provided a good introduction to numerical methods used in computer models.

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Appendix—Symbols and Definitions

a_j: coefficient of finite difference scheme

A: surface reflectivity (albedo)

b_j: coefficient of finite difference scheme

cld: cloud amount (in tenths)

c_j: coefficient of finite difference scheme

c_p: specific heat of air at constant pressure

d: index of agreement

d_j: coefficient of finite difference scheme

f: Coriolis parameter

F: function applied in solving for surface temperature through surface energy budget

g: acceleration caused by gravity

j: vertical level of model

k: Karman's constant (= 0.4)

K_h: eddy transfer coefficient for heat

K_m: eddy transfer coefficient for momentum

K_q: eddy transfer coefficient for moisture

l: mixing length

L_v: heat of transformation for water vapor

m: total number of iteration steps to solve for equation (10)

M_i: modeled data

n: total number of model or observed data points

N: total number of vertical grid points

O_i: observed data

p_j: coefficient in algorithm to solve for equation (14)

q: specific humidity

q_*: surface-layer turbulence scaling parameter for moisture

Q_s: soil heat flux

Ri: ratio of thermal to mechanical (wind shear) production turbulent energy called Richardson number

$\mathrm{Ri}_{\mathrm{crit}}$: limiting value of Richardson number

$R_{s\downarrow}$: incoming solar radiative flux

$R_{L\downarrow}$: incoming long-wave radiative flux

$R_{L\uparrow}$: outgoing long-wave radiative flux

s: local wind shear, $s = \sqrt{(\partial u/\partial z)^2 + \partial v/\partial z^2}$

t: time

T_c: temperature of cloud base (in Kelvin)

T_r: reference level (~2 m) temperature (in Kelvin)

u: east-west component of horizontal wind speed

u_g: east-west component of geostrophic wind speed

u_*: surface friction velocity

v: north-south component of horizontal wind speed

v_g: north-south component of geostrophic wind speed

w_{zi}: vertical velocity at z_i

W_*: vertical velocity scaling variable

x_j: coefficients in algorithm in equation (15)

y_j: coefficients in algorithm in equation (15)

z: height above ground level (agl)

z_i: height of top of planetary boundary layer

z_s: height of top of surface layer

β_τ: weighting function related to current timestep

$\beta_{\tau+1}$: weighting function related to future timestep

ϵ_c: emissivity of cloud base

ϕ_h: nondimensional temperature lapse rate

ϕ_m: nondimensional wind shear

ϕ_N: any profile variables (i.e., u, v, θ, or q)

θ: potential temperature

θ_*: potential temperature scaling constant

θ_{sfc}: surface potential temperature

θ_{2m}: potential temperature at 2 m

ρ: air density

σ: Stefan-Boltzmann constant

τ: indicates current timestep

$\tau + 1$: indicates future timestep

$\overline{\mathrm{any}}$: overbar denotes the mean

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13. ABSTRACT *(Maximum 200 words)*
This report examines details of a one-dimensional (1D) atmospheric boundary layer model to establish the proper functioning of its soil, plant, and atmospheric physics. To achieve this goal, I inspect, repair, and modify a computer program that scientists at the Hebrew University, Department of Soil and Water Sciences, gave to me years ago. The computer program was exercised to determine if the model results are stable when initial conditions are changed and to determine whether the results are sensible and generally consistent with observed data. To show this, I present a time series of the modeled surface energy budget and modeled profiles of boundary layer wind speed, potential temperature, and specific humidity for daytime (atmospherically unstable conditions) and for nighttime (atmospherically stable conditions). I compare these results, wherever practical, with observed meteorological data. From these results, I infer how well the transfers of momentum, heat, and moisture from one model layer to the next are characterized. I also present root mean square error and *d* values, where *d* is an index of agreement, to summarize the model results and comparison with observed data. From the results, I find that the 1D model is functioning properly in solving for many parameter relationships and is as reliable as the earlier models of this type in predicting the general features of boundary layer development.

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